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Surface Elevation Change and Mass Balance Of Icelandic Ice Caps Derived From Swath Mode CryoSat-2 Altimetry

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Key Points:

- Icelandic ice caps elevation change is the result of complex interaction between climate, ice dynamics and geothermal and magmatic processes
- Estimated rate of Icelandic mass loss between 2010 and 2015 is 5.8±0.7 Gt a^{-1} contributing 0.016±0.002 mm a^{-1} to sea level rise
- Swath processing improves on conventional POCA radar altimetry by providing elevation change with a two-fold increase in spatial coverage

Abstract

We apply swath processing to CryoSat-2 interferometric mode data acquired over the Icelandic ice caps to generate maps of rates of surface elevation change at 0.5 km postings. This high-resolution mapping reveals complex surface elevation changes in the region, related to climate, ice dynamics and sub-glacial geothermal and magmatic processes. We estimate rates of volume and mass change independently for the six major Icelandic ice caps, 90% of Iceland's permanent ice cover, for five glaciological years between October 2010 and September 2015. Annual mass balance is highly variable; during the 2014/15 glaciological year, the Vatnajökull ice cap (~70% of the glaciated area) experienced positive mass balance for the first time since 1992/93. Our results indicate that between glaciological years 2010/11and 2014/15 Icelandic ice caps have lost 5.8±0.7 Gt a^{-1} on average, ~40% less than the preceding 15 years, contributing 0.016±0.002 mm a^{-1} to sea level rise.

1 Introduction

It is estimated that glaciers and ice caps worldwide, including the periphery of the Greenland and Antarctic ice sheets, contribute about 47% of all land ice mass loss and 30% of current sea level rise [*IPCC*, 2013; *Gardner* et al., 2013]. Although satellite laser (LA) and radar (RA) altimetry observations have been crucial in estimating ice cap contributions to sea level change [*Bolch* et al., 2013; *Moholdt* et al., 2010a; *Moholdt* et al., 2010b; *Nuth* et al., 2010; *Rinne* et al., 2011a; *Rinne* et al., 2011b, *Gardner* et al., 2011; *Moholdt* et al., 2012; *Nilsson* et al., 2015a; *McMillan* et al., 2015], a comprehensive assessment is still lacking because of their complex topography, high slopes and small size with respect to satellite ground track spacing (7.5 km and 40 km at 60^{o}N for Icesat and ENVISAT, respectively) and footprint (2-10 km in diameter for ENVISAT). The European Space Agency (ESA) CryoSat-2 (CS2) satellite [*Wingham* et al., 2006] carries a state-of-the-art radar altimeter for land ice applications. CS2 improves upon previous missions in three ways: (1) narrow inter-track spacing (4km at 60°N) provides higher observation density, (2) synthetic aperture radar (SAR) processing along-track reduces the footprint size from ~1.65x1.65 km^2 (pulse limited) to ~1.65x0.305 km^2 (pulse-Doppler limited) and (3) the interferometer onboard CS2, in the so-called SARIn mode, allows the position of the surface reflection to be accurately located [*Wingham* et al., 2006]. Although these characteristics make standard CS2 SARIn elevations better suited to monitoring relatively small ice bodies characterized by complex and steep terrain [*McMillan* et al., 2014a; *Gray* et al., 2015], conventional Point-Of-Closest-Approach (POCA) altimetry tends to provide inhomogeneous spatial coverage due to the tendency of POCA towards sampling topographic highs (Fig. S4 and S6) [*Gray* et al., 2015].

Iceland is located at the boundary between polar and mid-latitude atmospheric circulation cells and between the warm Irminger and cold East Greenland / East Iceland oceanic currents. As a consequence, Icelandic ice caps are very sensitive to climatic shifts [e.g. *Björnsson* et al., 1998, 2013; *Aðalgeirsdóttir* et al., 2005; *Flowers* et al., 2005] and are estimated to have the highest static mass balance sensitivities among glaciers and ice caps north of 60° [*de Woul and Hock*, 2005]. They also display highly complex and dynamic behaviour unique to Iceland; about 60% of the current glaciated area lies over active volcanoes [*Björnsson and Pálsson*, 2008] and subglacial eruptions episodically trigger rapid ice loss albeit on short time scales (<1 year, [*Björnsson* et al., 2013]). Furthermore, surge-type outlet glaciers are present in all Icelandic ice caps, and cover 75% of Vatnajökull's surface [*Björnsson* et al., 2003]; surges in Iceland can cause significant mass transport to the ablation area and advance the terminus by up to 10 km during surge, with an opposite effect during multi-decadal post-surge periods [*Björnsson* et al., 2003; *Björnsson and Pálsson*, 2008; *Gourmelen* et al., 2011]. Furthermore, numerous active volcanoes lie below the ice surface so that subglacial eruptions episodically trigger rapid rates of ice loss albeit on short time scales (<1 year, [*Björnsson* et al., 2013]). Icelandic ice caps have been losing mass since the mid-1990s, in response to rising air temperatures caused by changes in atmospheric and oceanic circulation around Iceland, possibly induced by a global strengthening of the Atlantic meridional overturning circulation [*Björnsson* et al., 2013 and references therein]. Vatnajökull, with a loss of 6.58 Gt a^{-1} between 1995 and 2010, is the main contributor to the overall regional mass loss, followed by Langjökull (1.31 Gt a^{-1} between 1997 and 2010) and Hofsjökull (1.24 Gt a^{-1} between 1995 and 2010) in the central highlands (Table 1) [*Björnsson* et al., 2013]. Iceland as a whole has lost mass at a rate of ~11.0±1.5 Gt a^{-1} in the period 2003-2010 and contributed 0.03±0.004 mm a^{-1} to sea level rise [e.g. *Björnsson* et.al., 1998, 2002, 2013; *Guðmundsson* et al., 2011; *Jacob* et al., 2012; *Gardner* et al., 2013; *Pálsson* et.al., 2012; *Jóhannesson* et.al., 2013; *Hannesdóttir* et al. 2015; *Magnússon* et al., 2016; *Pope* et al 2016]. However, inter-annual variability is high, with rates of mass loss varying from 2 to 25 Gt a^{-1} between 1995 and 2009 [*Björnsson* et al., 2013]. This reflects both variability in tephra deposition on the ice caps [e.g. *Möller* et al., 2014] as well as their high sensitivity to temperature and precipitation [*Björnsson* et al., 2013, *Aðalgeirsdóttir* et al., 2006; *de Woul and Hock*, 2005].

Here, we extend mass balance estimates of the Icelandic ice caps from 2010 to 2015, by exploiting CS2 as a swath altimeter. We estimate the annual rate of mass change of Iceland's six largest ice caps, Vatnajökull, Langjökull, Hofsjökull, Mýrdalsjökull, Drangajökull and

Eyjafjallajökull, corresponding to 90% of the island's permanent ice cover, and over 99% of its volume [*Björnsson and Pálsson*, 2008].

2 Methods

We measure time dependent elevation over the ice caps using swath processing of CS2 level 1b SARIn data (SwSARIn). In contrast to the conventional POCA method, SwSARIn exploits the full radar waveform to provide a dense swath of elevation measurements across the satellite ground track (beyond POCA) when signal and surface conditions are favourable [see Supplementary Material; *Hawley* et al., 2009; *Gray* et al., 2013; *Christie* et al., 2016; *Ignéczi* et al., 2016]. As a reference, we also use elevations derived from the operational CS2 level 2 POCA product to assess ice cap elevation changes (see Supplementary Material), where POCA refers to the CS2 heights obtained via conventional retracking [*Wingham* et al., 2006]. For both datasets, we use CS2 baseline C data which are available from July 2010 to present.

We compute rates of surface elevation change $\dot{h}$ from SwSARIn data using a plane-fit algorithm [*McMillan* et al., 2014b] over five glaciological years: 2010/11 to 2014/15 (Fig. 1). We define one glaciological year as the period between 1st, October in year n and 30th September in year n+1. The dense elevation field provided by SwSARIn processing allows gridding at 0.5 km posting. In each pixel, the time dependent elevation is obtained by:

$$z(x, y, t) = c_0 x + c_1 y + \dot{h}t + c_2 \quad (1)$$

where x, y, t are easting, northing and time respectively. The time dependent coefficient retrieved from the model fit is the linear rate of surface elevation change, $\dot{h}$. The model is iteratively fitted to the data, excluding elevation differing from the model by more than 3 standard deviations, until no more outliers are detected. The pixel rate uncertainty $\varepsilon_{\dot{h}}$ is extracted from the covariance matrix of the model parameters (see Supplementary Material, Text S2). Pixels are discarded whenever a set of quality thresholds are exceeded (see Supplementary Material, Text S4) and final coverage of the rates of surface elevation change maps are 80% (Vatnajökull), 75% (Langjökull), 87% (Hofsjökull), 69% (Mýrdalsjökull), 65% (Drangajökull) and 27% (Eyjafjallajökull), respectively. No smoothing is applied, in order to minimize the correlation between adjacent measurements that would otherwise impact on the analysis of spatial variability in $\dot{h}$, and is permitted by the high observation density provided by SwSARIn.

We interpolate gaps in the maps of surface elevation change rates (Fig. 1) using hypsometric averaging [e.g. *Moholdt* et al., 2010a; *Nilsson* et al., 2015a] as a form of regionalization method and calculate ice cap volume changes from the gap filled maps (we do not use the method to extrapolate beyond the locus of the SWSARin measurements). We apply the regionalization independently for all of the ice caps except for Eyjafjallajökull which has relatively few measurements and is therefore processed together with the neighbouring Mýrdalsjökull. The resulting $\dot{h}$ map is divided into 50 m elevation bands using an external DEM from the National Land Survey of Iceland (Landmælingar Íslands, www.lmi.is) and the volume change $\dot{V}_k$ of each band k is calculated as the product of the mean $\dot{h}_k$ and the surface area A_k. The DEM spatial resolution is downscaled to the $\dot{h}$ grid resolution so that pixels elevations and elevation bands areas are representative of the pixels size. Volume change estimates for all bands are added together and then converted to a mass balance rate $\dot{M}$ using the density of glacial ice. Although this simplification ignores potential variations in snow/firn density (e.g. with elevation and thus melt), it is commonly used when deriving mass change and sea level contribution from ice caps [e.g. *Magnússon* et al., 2016, *Nilsson* et al., 2015a; *Nuth* et al., 2010; *Moholdt* et al., 2010b; *Björnsson* et al., 2013]. For comparison, we also provide a mass balance estimate assuming a dual density scenario [e.g. *Gardner* et al, 2011; *Moholdt* et al., 2010a] to account for density differences between the ablation and accumulation area. We propagate rate errors $\varepsilon_{\dot{h}}$ of the individual pixels to estimate uncertainties for $\dot{V}$ and $\dot{M}$ (see Supplementary Material).

3 Results

SwSARIn provides a step-change in surface coverage (Fig. 1), generating ~10 million elevation measurements over Vatnajökull between October 2010 and September 2015 and allowing the retrieval of rates of surface elevation change over 80% of the ice cap area (Fig. S5). In comparison, ICESat acquired 851 elevation measurements over all Icelandic ice caps between 2003 and 2009 [*Nilsson* et al., 2015a]. With the conventional POCA approach, CS2 delivers ~60,000 observations over Vatnajökull (October 2010 to September 2015) and provides rates of surface elevation change over 40% of the ice cap area, preferentially along topographic highs (see Text S3 in Supplementary Material and Fig. S4). Over the Langjökull ice cap, the particular hypsometry accentuates the concentration of elevations over the ice divide (inset in Fig.1, Fig. S6). There is almost no POCA observation close to the marginal areas of the northern dome (Fig. S8, centre panel) and only ~10 observation per km^2 over the southern dome (Fig. S8, right panel), which is insufficient to estimate robust rates of surface elevation change. In turn, limited sampling at the margins where most of the thinning is occurring impacts on the representativity of the POCA rates of volume and mass change (see Supplementary Material, Text S3).

[INSERT FIG. 1 AND CAPTION 1]

The time series of surface elevation change over the Vatnajökull ice cap (Fig. 2) shows a clear seasonal pattern with an increase in surface elevation during the accumulation period followed by a rapid decrease during the melt season, with amplitudes of about 3 m similar to observations over other Arctic ice caps [*Gray* et al., 2015]. Additionally, the elevation time series show an absence of sharp jumps in elevation that would otherwise be indicative of a sudden and unusual change in scattering horizon, and would introduce a bias in the estimated rates of surface elevation change [*Nilsson* et al., 2015b, *McMillan* et al., 2016].

[INSERT FIG. 2 AND CAPTION 2]

The data reveal a clear pattern of thinning, with rates of up to 10 m a^{-1} over most of the marginal areas of the ice caps, while change in the ice caps interior is more heterogeneous with both thinning and thickening observed (Fig. 1). This variability in the interior is particularly apparent over Vatnajökull, where several basins - e.g. Brúarjökull (Br), Síðujökull (Si), Dyngjujökull (Dy) - are thickening at high elevation while Skeiðarárjökull (east of Si) is thinning over almost its entire area. Thinning of Langjökull in the central highlands is widespread on the ice cap's surface up to, and including, the ice divide, while neighbouring Hofsjökull shows thickening over the centre and thinning over the margins. In the south of Iceland, relatively high rates of thickening (up to 3 m a^{-1}) are widespread over Mýrdalsjökull's central plateau. Thinning is visible particularly on its northern slopes which lie at low elevations as well as on the steeper southern margins. In the same region and despite being exposed to a similar climate, Eyjafjallajökull shows signs of thinning at its summit; however coverage here is limited due to the small area (~80 km^2) and steep hypsometry (~700-1560 m). Drangajökull (northwest) mostly displays a thickening pattern in the comparatively large accumulation area.

We use the CS2 derived rates of surface elevation change to compute mean annual rates of ice cap volume and mass change over five glaciological years from October 2010 to September 2015 (Table 1). During this time, we estimate that the Vatnajökull ice cap (~70% of Iceland's glaciated area) is losing mass at a rate of 3.68±0.61 Gt a^{-1} (-0.52±0.09 m_{we} a^{-1}) and is the main contributor (63%) to mass loss in Iceland, followed by Langjökull (12%) and Hofsjökull (8%) in central Iceland (Table 1). Langjökull is the fastest changing ice cap with -0.81±0.23 m_{we} a^{-1} specific mass balance, followed by Hofsjökull and Vatnajökull with -0.66±0.15 m_{we} a^{-1} and -0.52±0.09 m_{we} a^{-1}, respectively (Table 1). A combined estimate is generated for Mýrdalsjökull and Eyjafjallajökull (3.6% of loss) since data coverage over the latter is limited and the two ice caps are exposed to similar climatic conditions. To the northwest, Drangajökull appears to be close to balance (-0.05±0.07 Gt a^{-1}; -0.28±0.40 m_{we} a^{-1}); the uncertainty is comparatively large due to the small aerial extent and steep hypsometry of the ice cap (Table 1). Summing contributions from the six ice caps analyzed in this study, and rescaling for the remaining 10% glacierized area not included in our survey, we estimate that Iceland lost ice at a rate of 5.83±0.74 Gt a^{-1} (-0.59±0.07 m_{we} a^{-1}) between October 2010 and September 2015, corresponding to 0.016±0.002 mm a^{-1} eustatic sea level change. Assuming a dual density scenario in the ablation and accumulation areas with ρ_{abl}=900 kg m^{-3} and ρ_{acc}=650 kg m^{-3}, the mass loss and contribution to sea level change estimates are higher by just 4%, within the uncertainty of the single density case. During the glaciological year 2014/15, the Vatnajökull ice cap had positive mass balance (Fig. 2), an unprecedented observation in the last two decades [*Björnsson* et al., 2013] and due to anomalously high winter precipitation. This anomaly is reflected in the time series of surface elevation change where the trends in both the ablation and accumulation areas change after October 2014 (Fig. 2). In the 4 glaciological years before 2014/15, we find that Vatnajökull's rate of mass loss was 4.93±0.80 Gt a^{-1} (-0.69±0.11 m_{we} a^{-1}), or ~34% larger than the period 2010/11 to 2014/15.

We compared our geodetic estimates for the Langjökull ice cap and the Brúarjökull basin of the Vatnajökull ice cap against *in-situ* field derived mass balance observations from ongoing surveys [e.g. *Björnsson* et al., 1998, 2002, 2013; *Pálsson* et al., 2012; *Jóhannesson* et al., 2013]. We restricted the datasets to the same time period, four glaciological years from October 2010 to September 2014. The geodetic estimate for Langjökull, -0.76±0.25 Gt a^{-1} (-0.92±0.30 m_{we} a^{-1}), is 38% less negative than that from the *in-situ* data, -1.05±0.36 Gt a^{-1} (-1.28±0.30 m_{we} a^{-1}), but the two values agree within uncertainties. Over the Brúarjökull basin the agreement is good, -0.51±0.09 Gt a^{-1} (-0.37±0.07 m_{we} a^{-1}) compared to -0.49±0.22 Gt a^{-1} (-0.35±0.30 m_{we} a^{-1}) for the geodetic and *in-situ* values respectively. Using a dual density scenario, Langjökull's and Brúarjökull's geodetic mass balance estimates change by +17% and -18%, respectively.

4 Discussion

The heterogeneity of the rates of surface elevation change can be linked to the heterogeneity of ice caps hypsometry as well as their exposure to local climatic conditions, active volcanoes and glacier surge events. Individual basins of the Vatnajökull ice cap display distinct behaviours, either thinning across their entire length or experiencing thickening at high elevation. Three basins, namely Brúarjökull, Síðujökull and Dyngjujökull (Fig. 1), show large areas of thickening at higher elevation, as they are currently in a post-surge stage, responding to surges that occurred in 1963, 1995 and 1999, respectively [*Björnsson* et al., 2003; *Fischer* et al., 2003]. Thickening in the Gjálp area (Fig. 1), by an average of 0.7 m a^{-1}, is related to a combination of snow drift and ice inflow into the depression created by the 1996 subglacial volcanic eruption; these uplift rates are down from 40 m a^{-1} as measured in the year following the eruption [*Guðmundsson* et al., 2002]. North of Gjálp, over the Bárðarbunga central volcano caldera ice surface, the strong subsidence pattern is the surface response to the Bárðarbunga eruption that occurred between August 2014 and March 2015 [*Sigmundsson* et al., 2014; *Guðmundsson* et al., 2016]. This event deflated a magma chamber below the ~700 m thick ice; little or no ice was melted, but the caldera bedrock floor lowered by tens of metres and the ice above lowered similarly forming a cauldron like surface subsidence with a volume of ~1.9 km^3 [*Sigmundsson* et al., 2014; Guðmundsson et al. 2016]. The impact of this area on the ice cap wide rate of volume change is 0.05 km^3 a^{-1} (~1% of Vatnajökull's total volume change). In the central highlands, and despite their close proximity and similar climatic conditions, the pattern of rates of surface elevation change of the Langjökull and Hofsjökull ice caps differ considerably, most likely due to their differing hypsometry. Despite having similar area and volume (~900 km^2 and ~200 km^3), Langjökull has a lower elevation range (430-1440 m a.s.l) than Hofsjökull (620-1790 m a.s.l) [*Björnsson and Pálsson* 2008; *Guðmundsson* et al., 2009] and a large portion of the surface of Langjökull therefore lies close to the current equilibrium line altitude (ELA) [*Pálsson* et al., 2012]. Thickening is visible in the accumulation area of the West and East Hagafellsjökull basins of the Langjökull ice cap (Fig. 1) and is a dynamic response to the 1980 and 1999 surge events, respectively [*Björnsson* et al., 2003]. The central part of the Mýrdalsjökull ice cap is thickening at rates of about 1-3 m a^{-1} although the surface elevation of the plateau has not changed compared to 1999. The thickening is most likely induced by the extreme precipitation in winter 2015, which deposited 10-15 m of snow on the ice cap. Over Eyjafjallajökull's summit, the surface is thinning as ice flows into the crater created by the Eyjafjallajökull eruption in 2010 [*Oddsson* et al., 2016]. Over Drangajökull (northwest), despite the relatively small size of the ice cap as well as the steep elevation range, SwSARIn data captures the thinning pattern across the ablation area. This allows us to generate a robust estimate of mass balance, a result that cannot be achieved with conventional POCA processing (see Text S3 in Supplementary Material and Table S1 and S2).

Geodetic mass balance derived from repeat altimetry is dependent on the regionalization method chosen to derive volume change from the rates of surface elevation change [e.g. *Nilsson* et al., 2015a]. The high density of measurements provided by SwSARIn allows us to regionalize at the ice cap scale and in some cases at the basin scale (e.g. Brúarjökull), better accounting for local differences, in contrast to datasets with a lower density of observations which require mean hypsometric related rates of surface elevation change to be averaged at the scale of Iceland as a whole [*Nilsson* et al., 2015a]. Thus, the hypsometric averaging method applied at the basin scale shows good agreement with the *in-situ* estimate for one of Vatnajökull's largest basins: Brúarjökull. Comparing the SWSARIn and *in-situ* mass balance estimates over the Langjökull ice cap instead shows a difference between the two approaches. Current inter-drainage basin variability in rates of surface elevation change is relatively large in Iceland and is related to dynamic adjustment after glacier surges and sub-glacial eruptions as well as contrasting climatic conditions, e.g. due to inland precipitation shadow, hypsometry or distance from the south coast (the North-Atlantic low path). For example, the southeastern basins of Vatnajökull (e.g. as in Aðalgeirsdóttir et al. [2006]) reach low elevations at their termini, are exposed to high precipitation and have infrequent surges [*Björnsson* et al., 2003]. In contrast, basins in the northwest are more affected by surges and their termini are above 700 m elevation. Applying a hypsometric model at the ice cap scale would clearly not capture this complexity. SwSARIn provides a step change from previous altimetry-based techniques in mapping the complexity of ice caps' response to internal and external forcing as it enables the independent monitoring of individual ice caps. Additionally, the method can be used to derive mass balance estimates at the individual basin scale (e.g Brúarjökull).

5 Conclusions

CryoSat-2 swath radar interferometric altimetry (SwSARIn) increases the density of surface elevation measurements over Icelandic ice caps by 2 and 5 orders of magnitude with respect to the conventional point-of-closest-approach (POCA) method applied to the CryoSat-2 and ICESat missions, respectively. Compared to POCA measurements, which tend to concentrate on topographic highs, SwSARIn samples a wider range of elevations which helps generate more reliable estimates of mass balance, particularly for Icelandic ice caps with complex hypsometry. Swath altimetry allows high resolution mapping of surface elevation and its temporal change revealing complex spatio-temporal patterns of surface elevation change related to climatic, dynamic, and sub-glacial processes in Iceland. We estimate that Icelandic ice caps have lost a total of 5.8±0.7 Gt a^{-1} (-0.6±0.1 m_{we} a^{-1}) between October 2010 and September 2015, equivalent to 0.016±0.002 mm a^{-1} eustatic sea level change. This estimate suggests that over this 5 year period, the mass balance was 40% less negative than the preceding 15 years, a fact which partly reflects the anomalous positive balance year across Vatnajökull in 2014/15. Our observations also demonstrate the capability of SwSARIn elevations to image glaciological processes occurring at the sub-catchment scale, and to infer global, time-dependent, mass balance over region of complex hypsometry such as ice caps and ice sheet margins.

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Figures

Figure 1 - Rates of surface elevation change maps of Icelandic ice caps between 2010 and 2015 at 0.5 km posting based on SwSARIn heights as well as location of the ice caps in Iceland. V (Vatnajökull), L (Langjökull), H (Hofsjökull), M (Mýrdalsjökull), D (Drangajökull), E (Eyjafjallajökull). Basin outlines are shown in thin black lines. Selected basins over Vatnajökull and Langjökull are: Brúarjökull (Br), Síðujökull (Si), Dyngjujökull (Dy), Gjálp (Gj), Hagafellsjökull West (Hw) and Hagafellsjökull East (He) (thick black outlines). Ice caps areas after Björnsson and Pálsson [2008]. Contour elevations are shown in grey. The inset shows the location of individual elevation measurements using SwSARIn and POCA approaches over Langjökull.

Figure 2 - Vatnajökull elevation time series (60 days step) produced from SwSARIn elevations above and below 1200 m, used as an approximate ice cap wide ELA. Dark grey bands highlight the accumulation period between October and May; the non-shaded area corresponds to the ablation period between June and September. The two trends show mean rates of elevation change between 2010-2014 and between 2014-2016.

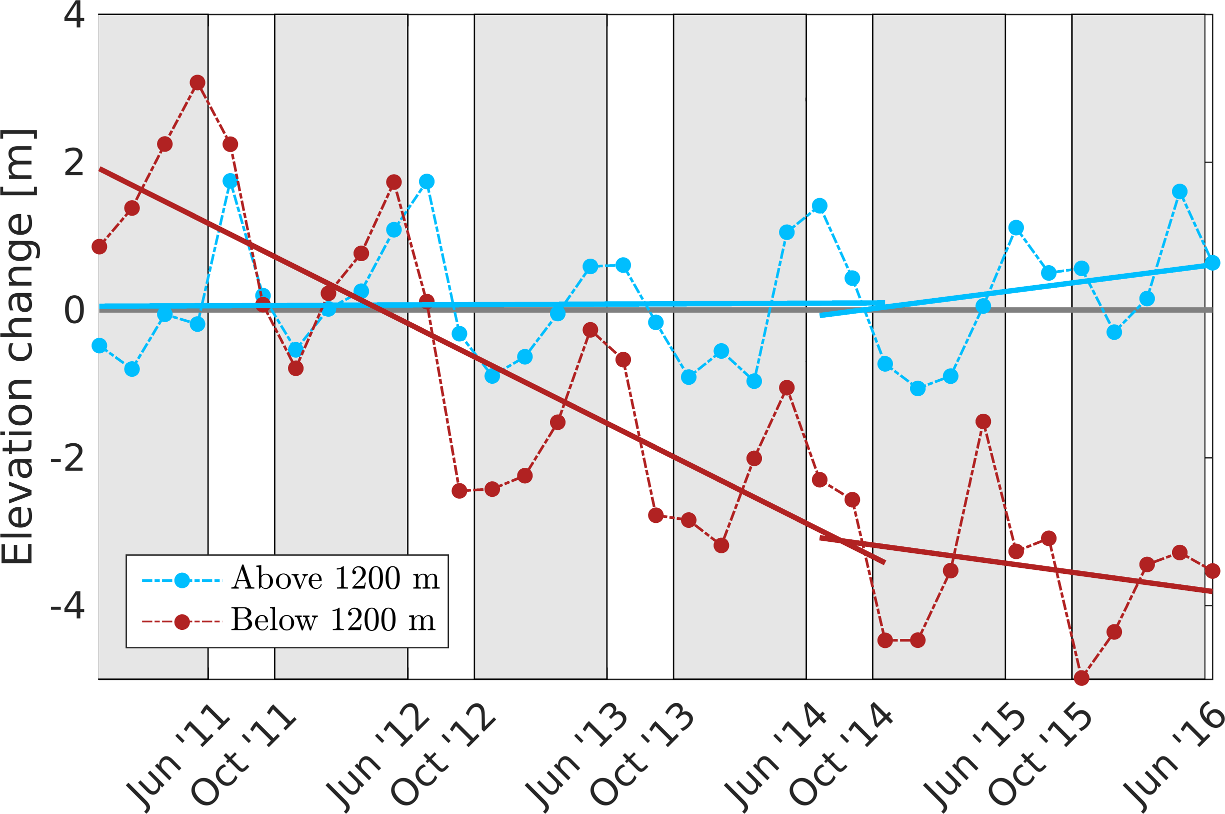
Tables

Table 1 – Mass balance of Icelandic ice caps

Estimates from SwSARIn data for five glaciological years between October 2010 and September 2015, as well as from the current literature (w.r.t. the specified time period). Mass change $\dot{M}$ is given in Gt a^{-1} as well as m_{we} a^{-1} (specific mass balance). Ice caps areas and volumes after *Björnsson and Pálsson* [2008]. References in table: (1) *Björnsson* et al. [2013], (2) *Jóhannesson* et al. [2013], (3) *Magnússon* et al. [2016], (4) *Guðmundsson* et al. [2008], (5) *Pálsson* et al. [2012], (6) *Gardner* et al. [2013], (7) *Jacob* et al. [2012] and (8) *Nilsson* et al. [2015a]. * Mass balance of Vatnajökull between October 2010 and September 2014 is -4.93±0.80 Gt a^{-1} (-0.69±0.11 m_{we} a^{-1}).

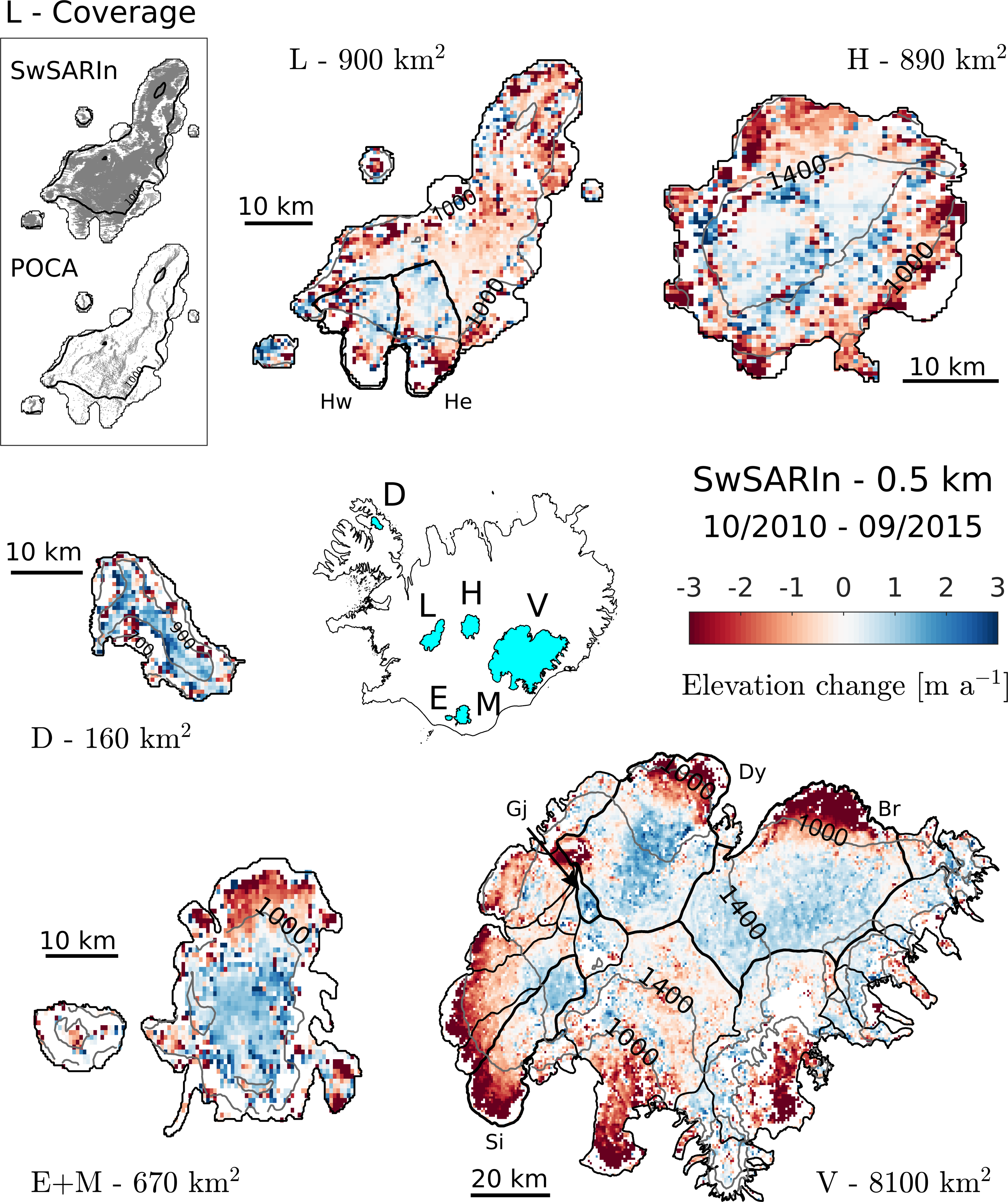
	A [km²]	**V [km³]**	**$\dot{M}$ [Gt a⁻¹] (period)**	**$\dot{M}$ [Gt a⁻¹]**	**$\dot{M}$ [m_{we} a⁻¹]**
Vatnajökull	8100	3100	-6.58 [1] (1995-2010)	-3.68* ± 0.61	-0.52* ± 0.09
Langjökull	900	190	-1.31 [1] (1997-2010) -0.91 [4] (1999-2007) -1.20 [5] (1999-2007)	-0.70 ± 0.20	-0.81 ± 0.23
Hofsjökull	890	200	-1.24 [1] (1995-2010) -0.92 [2] (2004-2008)	-0.45 ± 0.10	-0.66 ± 0.15
Mýrdalsjökull + **Eyafjallajökull**	590 + 80	140	-0.78 [4] (1999-2007) -0.06 [2] (2004-2010)	-0.21 ± 0.16	-0.39 ± 0.29
Drangajökull	160	24	-0.07 [3] (2005-2011) -0.05 [2] (1990-2011)	-0.05 ± 0.07	-0.28 ± 0.40
Iceland	~11,000	~3,600	-[9-11] ± [1-3] [1,6,7,8] (~1995-2010)	-5.83 ± 0.74	-0.59 ± 0.07

Figure 2.



4
2
0
-2
-4
Elevation change [m]
Above 1200 m
Below 1200 m
Jun '11
Oct '11
Jun '12
Oct '12
Jun '13
Oct '13
Jun '14
Oct '14
Jun '15
Oct '15
Jun '16

Figure 1.



L - Coverage
SwSARIn
POCA
L - 900 km^2
H - 890 km^2
10 km
1000
1400
Hw
He
D
L
H
V
E
M
SwSARIn - 0.5 km
10/2010 - 09/2015
-3 -2 -1 0 1 2 3
Elevation change [m a^{-1}]
D - 160 km^2
900
700
E+M - 670 km^2
20 km
Dy
Br
Gj
Si
V - 8100 km^2